Fast but spatially scattered smectite-formation in the proglacial area Morteratsch: An evaluation using GIS

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Fast but spatially scattered smectite-formation in the proglacial area Morteratsch: an evaluation using GIS

Markus Egli[a,*], Michael Wernli[b], Conradin Burga[a], Christof Kneisel[c], Christian Mavris[a], Giuseppe Valboa[d], Aldo Mirabella[d], Michael Plötze[e], Wilfried Haeberli[a]

[a]Department of Geography, University of Zürich, Winterthurerstrasse 190, 8057 Zürich, Switzerland

[b]Institute of Natural Resource Sciences, Zurich University of Applied Sciences, Grüental, 8820 Wädenswil, Switzerland

[c]Department of Physical Geography, University of Würzburg, Am Hubland, 97074 Würzburg, Germany

[d]Istituto Sperimentale per lo Studio e la Difesa del Suolo, Centro di ricerca per l'agrobiologia e la Pedologia CRA-ABP, Piazza D'Azeglio 30, 50121 Firenze, Italy

[e]Institute for Geotechnical Engineering, ETH Zurich, CH-8093 Zurich, Switzerland

*corresponding author:

tel. +41 44 635 51 14

fax: +41 44 635 68 48

e-mail: megli@geo.unizh.ch

Abstract

Proglacial areas in the Alps usually cover a time span of deglaciation of about 150 years (time since the end of the ‘Little Ice Age’ in the 1850s). In these proglacial areas soils have started to develop. Due to the continuous retreat of the Morteratsch glacier (Swiss Alps), the corresponding proglacial area offers a continuous time sequence from 0 to 150 year-old surfaces. Furthermore, an optimal digital dataset about the development of vegetation and soils is available for this area. The soils have been developing on glacial till having a granitoidic character. We investigated the clay mineral assemblage at 35 sites within the glacier forefield. Smectite can be used as a proxy for weathering intensity in these environments. In the proglacial area, the smectite concentration in the topsoil steadily increased with time of weathering; however, this development displayed a strongly scattered trend. The complex interplay of biological, physical, and chemical processes in pedogenesis and clay mineral evolution limits our ability to predict and interpret landscape dynamics. We consequently tried to analyse the role of topographic and vegetation modifications on the smectite content. Sites not or only slightly prone to erosion (flattening slope ridge, steepening slope ridge) or flat morphology promoted the formation of smectite. In addition, the texture of the soil material influenced soil moisture and hence the degree of weathering and the development of vegetation. Although vegetation is not a fully independent factor, because it is interrelated to the surrounding environmental conditions, it seemed to exert its influence on weathering and, consequently, the formation of smectite. Green alder stands and grass heath, where moister soils develop that have a finer texture or where more organic acids are produced, were correlated with a higher smectite content. Humus forms serve as a proxy for the soil biota and soil organic matter composition. At sites having a Eumoder and a higher soil organic matter content, the smectite concentration was elevated. At these sites, the production of chelating compounds or organic acids in the soil is believed to promote the development of smectites via an intermediate stage of hydroxy-interlayered minerals and the subsequent removal of the hydroxide polymers. In this work we have demonstrated that the topographic signature and the effect of vegetation on the formation of smectite and consequent weathering are pervasive. Our results will serve as a basis for further spatio-temporal modelling.

Keywords: smectite, proglacial area, temporal evolution, topography, vegetation, green alder, humus form

Introduction

Weathering in cold regions has often focused on the notion of ‘cold’. As a result of this approach, the predominating perception in terms of weathering has been that mechanical processes predominate, with freeze-thaw weathering as the prime agent and that chemical processes are temperature-inhibited, often to the point of non-occurrence. Indeed, many cold regions show similar or even more intense weathering assemblages than those in warmer regions (e.g. Hall et al., 2002; Föllmi et al., 2009a,b). Contrary to popular presentations, several investigations meanwhile document that weathering in cold Alpine regions, including chemical weathering, is not strictly temperature-limited but rather controlled by moisture availability (Egli et al., 2006). Furthermore, the availability of fresh mineral surfaces that are provided by physical erosion determines chemical weathering rates. Glaciers and periods of glaciation may have a significant impact on global weathering, changing the interplay between physical and chemical weathering processes by putting large volumes of dilute meltwaters and fine-grained sediments in contact with each other (Föllmi et al., 2009a,b; Arn et al., 2003). Glaciers are significant agents of physical erosion; for example, the mechanical denudation of glacierised valleys in Alaska, Norway and the Alps is an order of magnitude greater than that in equivalent non-glacierised basins (Hallet al., 1996; Föllmi et al., 2009b). Chemical weathering intensities depend mainly on the lithology (e.g. highly reactive minerals such as carbonates and sulphates vs. crystalline rocks), the accumulation of organic matter (Conen et al., 2007), the rate of supply of fresh rock material, the age of exposure, and the character of the hydrological drainage system. In cryic, ice-free environments, chemical weathering can be an active process leading to the formation of clay minerals (e.g. Simas et al., 2006; He and Tang, 2008). In arctic tundra and partially also in Alpine regions, soil formation and weathering processes may be spatially highly patterned. Cryoturbation and other cryogenic processes produce stripes, non-sorted polygons, earth hummocks etc. Cryogenesis is a controlling factor in patterned ground formation resulting in cryoturbated soil profiles, cryostructures and carbon sequestration (Ugolini, 1986; Haugland, 2004; Ping et al., 2008). Frost disturbance interrupts soil-forming processes and produces a time-lag effect of pedogenesis at such sites compared to soils that are not affected by permafrost (Haugland, 2004).

With respect to chemical weathering, several studies have shown that smectitic constituents can be taken as an indicator of weathering intensity in Alpine soils and cold regions (Righi et al., 1999; Egli et al., 2001a,b, 2003, 2006). Investigations of Egli et al. (2001a) have revealed that the greatest changes in soil chemistry of Alpine soils on granitic host material occur within the first 3000 – 4000 years of soil development. Element leaching rapidly slows down after this period. Extensive mineral weathering usually results in significant losses of chlorite and mica and the corresponding formation of smectite and regularly and irregularly interstratified mica-smectite. Highest weathering and transformation rates of clay minerals should be measured at the initial stage of soil formation (Egli et al., 2003a; Mavris et al., 2010). Similarly, Hosein et al. (2004) and Föllmi et al. (2009a,b) have shown that biotite weathering in young soils was generally much higher than in old soils (11000 years). Calculated weathering rates of biotite were several orders of magnitude higher than known field weathering rates. This seems to be due to the predominance of fine-grained (< 63 μm) particles in glacial sediments that are mechanically disaggregated and preferentially leached. Proglacial environments are important for the understanding of global CO_2 cycling on glacial/interglacial timescales as they made up a significant amount of the global land surface during the Quaternary due to the advance and retreat of glaciers and ice sheets (Gibbs and Kump, 1994).

Soils in proglacial areas in the Alps are young and have been forming over a time span of about 150 years (the time since the end of the 'Little Ice Age' in the 1850s; Fitze, 1982). The proglacial areas are in most cases defined as the zone between the present-day glacier and the terminal moraines deposited in the 1850s. The aim of our research was to examine whether changes of the smectite content along a temporal gradient of very young and consequently reactive soils can be determined. Furthermore, our aim was to relate the concentrations and changes of clay minerals not only to the soil-forming factor time but also other extrinsic factors such as topography and vegetation (Jenny, 1941) with the help of geographic digital datasets.

Investigation area

The soils studied lie within the proglacial area of Morteratsch in the Upper Engadine (Switzerland). The border of the proglacial area is given by terminal moraines deposited in the 1850s during the ‘Little Ice Age’ (Figs. 1 and 2). The actual length of this proglacial area is approx. 2 km and has an area of 1.8 km^2. The proglacial area is in a valley that runs N-S. The altitude ranges from 1900 m asl to about 2100 m asl. The backwall of the glacier is formed by the high mountain peaks of the South Raetian Alps such as Piz Bernina (4049 m asl, the highest peak of the Grisons and the Eastern Alps) and Piz Palü (3901 m asl). Alpine glaciers have fluctuated during the last 12000 years near the terminal moraines formed in the year 1850 indicating more or less similar climatic (having mean annual temperature fluctuations in the range of 1 – 1.5 °C) and hydrologic conditions within that period. This has been shown by many geomorphologic and climatic studies (cf. Burga and Perret, 1998; Keller, 1994; Magny, 1992; Maisch, 1992; Renner, 1982; Gamper, 1985; Patzelt, 1977). The glacial till in the glacier forefield consists of granitic and gneissic material and was produced through glacial transport within a small area of relatively homogeneous parent material. The lithostratigraphic units are mainly the Bernina- and Stretta-crystalline (Büchi, 1994; Spillmann, 1993). The parent material was greenschist facies overprinted (Trommsdorff and Dietrich, 1999). Present climatic conditions for the Morteratsch site are approx. 0.5 °C mean annual air temperature and approx. 1000-1300 mm mean annual precipitation as calculated by using data from the nearby meteorological stations Samedan and Bernina.

Materials and methods

Digital data basis and procedure

Soil map:

The soil map has a scale of 1:10000 (Egli et al., 2006; Fig. 2). Soil cartography and classification were performed according to the FAL (Swiss soil classification system; Brunner et al., 1997). Soil units were translated into the WRB system (IUSS working group, 2006). Soil mapping already includes a certain generalisation as small-scale variations (approx. < $100m^2$) could not be considered. The map items included soil type, soil depth (per unit area) relevant for plant growth (= soil volume – skeleton volume – groundwater volume; result is related to depth instead of volume), parent material, vegetation, topography, soil hydrology, terrain form, pH-value, soil organic matter content, soil skeleton (material > 2 mm as a whole), gravels and stones (gravelly: predominantly material < 5 cm in diameter; stony: predominantly material > 5 cm in diameter), granulometry, soil structure and humus form. The soils vary from Lithic Leptosols to Dystric Cambisols according to the World Reference Base for Soil Resources (WRB) soil classification (IUSS working group, 2006). In total, 47 soil pits were made and described in detail and used as reference profiles for mapping. The number of profiles varied according to the area of the soil units: 1 for Endogleyic Cambisol, 3 for Dystric Cambisol, 18 for Humi-Skeletic Leptosol (including Ranker), 3 for Haplic Fluvisol Endoskeletic and 22 for Skeletic/Lithic Leptosol. Ten pits (that are related to a botanical monitoring; Burga, 1999) were excavated for the chemical and physical analysis of the soil material.

Vegetation map:

The first detailed vegetation map (scale 1:10000) of the glacier forefield Morteratsch was established by Bäumler (1988). A slightly generalised map on a smaller scale (1:13700) of the actual vegetation cover (modified and completed after Fischer, 1999) gives an overview of the vegetation pattern in the glacier foreland (Fig. 2). Generally, three plant species groups were distinguished (Burga, 1999): a) pioneer species, starting early in the chronosequence and reaching their optimum in early or medium stages, b) subalpine forest and dwarf-shrub/grass heath species – most of them starting in later stages, c) species occurring mainly in alpine grassland having different distributions and optima along the gradient. Burga (1999) established ten permanent plots within the glacier foreland to monitor the vegetation dynamics of deglaciated areas. With the help of the vegetation cover of permanent plots and the time since ice retreat (chronosequence 1857 – 1997), a generalised model of primary plant succession could be established (Burga et al., 2010).

Soil sampling and mineral analyses

With respect to the clay mineralogy, surface soil samples (topsoils) were collected from 35 soil pits distributed over the whole proglacial area (Fig. 1) forming a soil chronosequence ranging from 0 to 150 yr. For each sample, approximately 1 – 2 kg of material was collected. Eleven of these sites were analysed in detail for both soil chemistry and mineralogy (Mavris et al., 2010). At the other 24 sites, only the mineralogical composition was measured in detail. Soil chemical data of additional 10 soil pits (cf. Mavris et al., 2010) also served as a basis for the soil map (see above), For the GIS analyses (see below) soil acidity, texture, soil organic matter, soil thickness and humus form were retrieved from the soil map.

The soil clay fraction (< 2 µm) of all 35 sites was obtained after destruction of organic matter with dilute and Na-acetate buffered H_2O_2 (pH 5) by dispersion with Calgon and sedimentation in water (Jackson, 1956). Oriented specimens on glass slides were analysed by X-ray diffraction using Cu-$K\alpha$ radiation (3-kW Rigaku D/MAX III C diffractometer, equipped with a horizontal goniometer and a graphite monochromator) from 2 to 15°2θ with steps of 0.02°2θ at 2 s/step. The following treatments were performed: Mg saturation, ethylene glycol solvation (EG) and K saturation. The K-saturated samples were, after XRD measurement at room temperature, heated for 2 h at 335 and 550 °C.

When chlorite was present, the presence of kaolinite was checked with FT-IR (Brooker Optics, Tensor 27) analysis (OH-stretching region near 3695 cm^{-1}) on powder samples (3%

ground soil material, 97% KBr) heated at 80 °C for ≥ 2 h. Digitised X-ray data were smoothed and corrected for Lorentz and polarisation factors (Moore and Reynolds, 1997). Diffraction patterns were smoothed by a Fourier transform function and fitted by the Origin™ PFM program using the Pearson VII algorithm. Background values were calculated by means of a non-linear function (polynomial 2nd order function; Lanson, 1997).

The semi-quantitative estimation of phyllosilicate concentration was performed by the combination of the areas of the ethylene glycol solvated, the Mg-saturated, the K-saturated and the heated (335 °C and 550 °C) samples. On the basis of the obtained integrals, an estimate of clay minerals composition was performed. The sum of the areas between 2 and 15°2θ, which were attributed to HIV (hydroxy interlayered vermiculites), smectite, vermiculite, mica, chlorite and kaolinite, were normalised to 100%. Although the (semi-)quantification of clay minerals in soils is plagued by manifold problems (Kahle et al., 2002), the applied and standardised (sample preparation, treatments, measurement and calculation) procedure enabled the assessment of the variability of clay mineral assemblage amongst the sites.

Area statistics

Area calculations (proportion of different soil types between two isochrones and in relation to either time, topographic or vegetational features) and statistics (regression analyses) were performed using ArcGIS 9.3 (ESRI) with modules programmed in Visual Basic for Applications (VBA). Input data sets were the digital soil and vegetation maps, the glacial states (Burga, 1999) and the digital elevation model (raster of 20 m) within the proglacial area. The calculations were done raster based (GRID, 20 m resolution).

The landscape forms (Tables 1 and 2), slope (Table 3) angle and exposure were related to the clay mineral composition. The different exposures were defined as follows: North (315 – 45 °N), south (135 – 225 °N), east (45 – 135 °N) and west (225 – 315 °N). Furthermore (and similar to a previous work; Egli et al., 2006), the exposures were simplified to north and south with north-facing defined as > 270 – 90 °N and south-facing as > 90 – 270 °N. Relative area calculations refer to the area between two isochrones of deglaciation (Figs 1 and 2). In total, 11 different glacial states (with corresponding isochrones) could be distinguished (Burga, 1999).

Following the suggestions of Winkler (2000), we used a standard error based on the standard deviation in a 95% confidence interval to get statistically significant differences between the data populations:

$$x \pm 1.96 \times \sqrt{\left(\sigma / \sqrt{n}\right)} \quad (1)$$

where x is the arithmetic mean, σ the standard deviation and n corresponds to the number of measurements.

The clay minerals data distribution (whole dataset as well as dataset of individual classes; see the results chapter) was checked for normal distribution using a Shapiro-Wilk test (SigmaPlot 11.0). Almost all datasets (whole datasets as well as datasets within the individual classes) were not significantly different from a normal distribution. This justifies the above-stated approach of using the standard error and confidence interval.

Results

Vegetation and soil development

The different vegetation units of the proglacial area are given in Table 4. Primary plant succession of the pro-glacial area with its predominantly siliceous parent material starts with the pioneer plant communities Oxyrietum digynae and Epilobietum fleischeri and ends with the larch/Swiss stone pine forest (Larici-Pinetum cembrae) after substantially more than 150 years. The first plants, i.e. *Epilobium fleischeri, Oxyria digyna, Linaria alpina*, *Saxifraga aizoides*, *Rumex scutatus*, appear about 7 years after deglaciation and reach greater cover-abundance values (about 5 – 25%) after about 27 years. The first species of the short-living Oxyrietum digynae appears approx. 10 years after deglaciation and disappears approx. 30 years later. The first small larch trees, the first shrubs of willows and green alder and the first dwarf-shrubs (e.g. the rust-leaved alpenrose) are found on areas that have been ice-free for about 12 to 15 years (Fig. 2). The establishment of *Larici-Pinetum cembrae* takes place after about 77 years. Surprisingly, lichens such as *Stereocaulon* cf. *dactylophyllum* need about 15 – 20 years to establish their first populations (Burga et al., 2010).

The dominant soil units (some sites do not have a soil) in the proglacial area are Haplic Fluvisols Endoskeletic, Skeletic or Lithic Leptosols, Humi-skeletic Leptosols (Fig. 2), including some sites with Ranker (IUSS working group, 2006), that have a weak B horizon and Dystric and Endogleyic Cambisols (endoskeletic)). The young soils that are close to the glacier showed almost no morphologic signs of chemical weathering and alteration products. They were usually characterised by a very thin and often discontinuous humus layer (Fig. 2). The oldest soils (150 years) often had a spatially continuous layer (O or A horizon) with organic matter and some signs of weathering-product formation (formation of Fe- and Al-oxyhydroxides, start of clay mineral formation/transformation) and, thus, a weakly developed B horizon (data shown in Table 2 of Mavris et al., 2010).

Smectite formation

To give an impression of the clay development in the proglacial area, the X-ray diffraction patterns of a soil clay material in direct proximity to the glacier and a surface soil sample having an age of 140 years are compared (Fig. 3). The XRD pattern of the material that was very recently exposed to weathering (t = 0) showed distinct and sharp reflections at 1.42, 1.23, 0.99 and 0.71 nm. Furthermore a very weak peak near 1.64 nm could be observed after ethylene glycol (EG) solvation. The peak at 0.85 nm can be attributed to amphibole and the one at 0.64 nm to plagioclase. Together with the other diagnostic treatments (K-saturation, heating at 335 °C and 550 °C), the clay minerals mica, chlorite, interstratified mica-HIV and traces of an expandable mineral (smectite) could be discerned. After 140 years, the XRD pattern of the clay fraction slightly changes. The peak at 1.67 nm (smectite) after EG solvation is more pronounced. Furthermore, a broadening of the 1.00 nm peak (with another peak near 1.08 nm) could be detected showing that mica is weathering. All other clay phyllosilicates were also detected. It seems that the concentration of chlorite also decreased slightly (Fig. 3; heating at 550 °C).

In principle, a significant ($p = 0.033$) trend could be observed along the investigated chronosequence (Fig. 4): the content of smectite increased with increasing exposure age of the substrate. Traces of smectite could already be detected in the parent material. The variability of the smectite concentration along this sequence was, however, considerable.

Mineral weathering started immediately at the beginning of soil formation and resulted in a significant loss of mica and a corresponding formation of smectite (Fig. 5). Chlorite also exhibits a decreasing (but statistically not significant) tendency when compared to smectite. Although the soils were very young and the concentration of smectite was rather low, measurable changes occurred within a very short period of time.

Effect of topographic and vegetational features on smectite formation

The patterned distribution of the soils (especially at their earliest stage) suggests that the soil-forming conditions were not identical everywhere. On a larger scale, the mineralogy and chemistry of the parent material in the proglacial area can be considered to be more or less constant. In this case, the parent material can, thus, be supposed to have a negligible influence as pedogenic factor (Jenny, 1941) in the differentiation of the soils across the selected sites. Also climate does not vary greatly in the area of interest and can therefore be considered a negligible factor.

The vegetation seemed to influence the development of smectite and other phyllosilicates (Fig. 6). The lowest smectite content was measured on sites without vegetation. A measureable, but statistically insignificant, increase in the smectite content was evident over the entire range from the pioneer grass communities to the Epilobietum fleischeri having single willow shrubs and alpenrose and finally to the green alder scrub (Alnetum viridis). Rock vegetation and boulder plant communities resulted in low smectite concentrations. The number of observations (smectite) in these units was, however, very low. Similar to the green alder scrub, grass heath on moister soils seemed to promote smectite formation. But also here, the number of observations (n ≤ 2) was too low for reasonable statistics. The smectite content was not measured in areas having the vegetation type 9 and 12 (Table 4).

Higher smectite content correlated with a higher OM content, while for chlorite and mica an opposite trend could be observed. Consequently, the amount of soil organic matter (OM) significantly influenced the clay mineral assemblage (Fig. 7). Not only soil organic matter *per se* exerted an influence on smectite: also the humus form significantly determined the formation of smectite (Fig. 7). Eumoder seemed to enhance the smectite formation compared to initial moder or no clear humus structures. Transitional weathering products, such as mica-HIV, also seem to be influenced by the humus form (although not significantly).

The concentration of smectite at north-facing (as well as west- and east-facing) sites was slightly higher than at south-facing sites (Fig. 8). For chlorite, an opposite trend (which would make sense in terms of weathering) was measured. If a subdivision of the exposure is undertaken only into north- (> 270 – 90 °N) and south-facing (> 90 – 270 °N) sites, the differences are less obvious (data not shown). A slightly lower smectite content and a higher chlorite content at south-facing sites can still be measured.

The form of the landscape should also influence soil formation (see Egli et al., 2006). As shown in Figure 8, it can be supposed that the clay mineral assemblage was to a certain degree determined by the landform. Sites having a high susceptibility to erosion showed higher mica and the lowest smectite concentrations.

The influence of the slope angle on the clay mineral assemblage is low. There is a tendency (but barely significant) to higher smectite and mica-HIV contents with decreasing slope angle (Fig. 8).

Physical properties of the soil material were correlated to the clay mineral assemblage. Sites having a finer texture (loamy sand; Fig. 9) showed a higher smectite content than sites having a sandy or gravelly texture. With time, soil thickness also increased due to weathering, unless a high erosion activity was present. The amount of smectite increased as the thickness of the soil increased (Fig. 9).

Discussion

Proglacial environments and areas having thawing permafrost are important for the understanding of global CO_2 cycling on glacial/interglacial timescales as they made up a significant amount of the global land surface during the Quaternary due to the advance and retreat of glaciers and ice sheets (Gibbs and Kump, 1994; Keller et al., 2007). These areas are often zones of high geochemical reactivity due to the availability of freshly ground reactive material or almost unaltered rock fragments, high water:rock ratios and contact times, high permeability and constant supply of dilute waters to percolate through the deposits (Föllmi et al. 2009 a,b). Chemical weathering is ultimately coupled to mineral formation and transformation processes. The weathering of silicates and formation or transformation of minerals theoretically depend on mineral reactivity, the supply of minerals, water, acid reactants, ligands and the Arrhenius rate law (Lasaga et al., 1994; West et al., 2005).

Soil development proceeded quickly in the proglacial area. Within 150 years, significant soil formation and clay mineral transformations took place. As hypothesised by Egli et al. (2003a), the transformation rate of mica (and probably chlorite) at the beginning of soil formation is particularly high. The formation of smectite is predominantly due to the transformation of precursor products such as mica (Fig. 5), chlorite (Righi et al., 1993; Carnicelli et al., 1997; Egli et al., 2001a,b) or, to a lesser extent, amphiboles (Dreher and Niederbudde, 2000; Mavris et al., 2010). With time, the smectite content in the clay fraction continuously increased at the expense of mica and, secondarily, chlorite (Fig. 5). This increase was, however, highly scattered. Apart from the above-mentioned transformation processes, patterned structures may be associated with abrupt thresholds that either enhance or stop/hinder soil formation and mineral transformation. This might be due to several causes such as microclimatic conditions, micro-relief (both parameters were not any of into account in this investigation), deposition of physically inhomogeneous parent material (sites with a more fine-grained deposit close to rock debris) and probably also to brief periglacial activity (cf. Haugland, 2004; Egli et al., 2006).

Besides the factor time, topographic characteristics, vegetational features and probably physical characteristics of the parent material may consequently also be responsible for the variability of smectite formation along the chronosequence. According to the paradigm of Jenny (1941), the factor vegetation is not a fully independent one. The smectite content that can be taken as an indicator of weathering intensity (Egli et al., 2003a,b) could also be related to factors other than only time: other influences on the spatial pattern of smectite could be the vegetation distribution, soil organic matter, humus forms, exposure, relief forms and grain size.

Egli et al. (2006) showed that weathering differed between south- and north-facing sites with north-facing sites having the higher weathering intensity. Usually, north-facing sites are hypothesised to have moister soils that give rise to a more intense weathering regime. Our results indicate that the exposure probably also has some effects on the clay mineralogy (Fig. 8). The results of Roering et al. (2010) suggest a significant biological role in sculpting landscapes and driving weathering and physical conversion of bedrock to soil. The properties of the substratum such as the texture of the parent material, the distribution of moisture and the availability of nutrients are essential factors for the establishment of the vegetation (Burga, 1999). Small-scale factors such as the texture determine plant growth and, in a further step, weathering and soil evolution (Burga et al., 2010). A higher content of smectite was found in soils having the finest texture in the proglacial area (sandy loam). A finer texture also retains more water and makes it available for weathering reactions. Green alder stands and grass heath on moister soils were correlated to the presence of Humi-skeletic Leptosols (the most developed soils in the proglacial area) and consequently to a higher content of smectite. Once vegetation is established, it influences weathering processes by releasing organic ligands and acids that promote dissolution and transformation reactions (Caner et al., 2010). Grass heath is known to be an important source of organic acids (Matzner and Ulrich, 1980) that finally may promote the formation of smectite. Furthermore, grass heath is found at rather moist sites. The higher availability of water finally also promotes dissolution and transformation reactions. In contrast, rock vegetation and boulder plant communities develop on a strongly gravelly (diameter < 5 cm; 20 – 30 vol.%) and strongly stony (diameter > 5 cm; 20 – 30% vol.%) substrate (Burga et al., 2010). The predominance of gravel and stones in the composition of the material leads to low smectite concentrations.

Humus forms and soil organic matter closely relate to the vegetation, the soil biota and climate. This interrelationship is furthermore based on the fact that humus forms result from the activity of soil organisms and, at the same time, act as habitat for them (Lalanne et al., 2008). The composition of plant litter greatly influences microbial degradation and, therefore, also SOM stability (Williams and Gray, 1974; Zunino et al., 1982). Humification is related to the preferential oxidation of plant polysaccharides, the selective preservation of more recalcitrant organic compounds such as lignin and phenolic structures and to the incorporation of organic compounds of microbial origin (Zech et al., 1997; Rosa et al., 2005). The humus forms serve as a proxy for the soil biota (macro- and micro-biology) and soil organic matter composition. The morphological properties of humus forms are known to correlate with soil biodiversity and thus with the decomposition rate of organic material as a result of soil biological activity (Klinka et al., 1990; Broll, 1998; Ponge et al., 2002; Sartori et al., 2005). The highest accumulation of soil organic matter in the proglacial area, the highest acidity and the highest production of organic ligands must be expected in the humus form Eumoder (Zanella et al., 2009). Eluviation of chemical elements was highest at the sites having a Eumoder where the highest amount of smectite was also measured. Organic ligands enhanced weathering (leaching of elements) and promoted the formation of low charge expandable minerals (smectites; Egli et al., 2008a). The Eumoder humus form, representing acidic conditions and conditions where organic ligands are formed, is a transition from a Hemimoder to a Dysmoder (Green et al., 1993; Zanella et al., 2009).

Clay minerals and other weathering products contribute to the stabilisation of soil organic matter. As shown by Denef and Six (2004), Egli et al. (2008b) and other authors, the type of clay minerals can affect the macroaggregate formation and stabilisation (and consequently organic C). This may be coupled to feedback regarding the formation of clay minerals.

High precipitation rates and the production of chelating compounds in the soil are believed to accelerate the formation of smectites (that in these environments are the weathering end-products of the phyllosilicates mica or chlorite) through the intermediate stage of hydroxy-interlayered minerals and the subsequent removal of the hydroxide polymers (Malcolm et al., 1969; Senkayi et al., 1981; Righi et al., 1999; Carnicelli et al., 1997; Egli et al., 2003b).

According to Anderson et al. (2000), silicate-weathering reactions are substantial only after vegetation cover is established. In proglacial areas, however, fresh mineral surfaces are provided by physical erosion of the parent material by the glacier. Freshly-ground material is very reactive and gives rise to high chemical weathering and mineral transformation rates (Föllmi et al., 2009a,b). Geochemical and physical controls of weathering fluxes are interrelated (Anderson et al., 2007; Riebe et al., 2004; Heimsath et al., 2001). Jin et al. (2010) found that chemical weathering was higher in upslope areas and slowed down in the valley floor. This seems also to be true, at least partially, for our investigation area. Smectite concentrations were highest in areas having a potentially slight erosion activity and on flat slopes. Lowest concentrations were measured in the valley floor. Sites on a flat slope (with an assumed more-or-less balanced in- and output of surface material), sites having a low susceptibility to erosion and sites having a high susceptibility to accumulating surface material showed the most advanced weathering stages with the highest smectite concentrations (Fig. 8). Our findings also agree well with the rather large-scale, 'macroscopic' investigations. Egli et al. (2006) found that better-developed and thicker soils were found at sites having a tendency to accumulation (such as depressions and at the foot of the slopes) as well as on concave slopes or flat slopes. At some sites, furthermore, fluvial erosion was a mechanism causing lower smectite content in the valley floor of the Morteratsch proglacial area. Although high erosion rates are coupled with high chemical weathering rates (Heimsath et al., 2001; Riebe et al., 2004; Anderson et al., 2007; Jin et al., 2010), this does not necessarily mean that highly weathered soils and consequently a high amount of smectite are always encountered at such sites. In fact, weathering products (newly formed or transformed mineral species) might be eroded and, consequently, weakly weathered or freshly-ground primary minerals may accumulate.

Conclusion

We studied the scattered formation of smectite along a chronosequence having a time span of 150 years. The following were the principal findings:

- Smectite was actively formed in the proglacial area but it exhibited a strongly scattered trend. Elevated smectite contents were found at sites having a thicker soil.
- Topography pertains to the configurations of the landscape and refers to inclination, concavity or convexity and (north- and south-) exposure. Sites showing slight erosion (flattening slope ridge, steepening slope ridge) or flat slopes had the highest smectite content. Sites having high erosion rates and valley floors (where either accumulation may occur or erosion due to fluvial processes) exhibited a lower smectite amount.
- The texture of the soil material influenced weathering and development of vegetation. A higher content of smectite was found in the finest-textured material of the proglacial area (sandy loam).
- Vegetation cannot be considered as a fully independent factor as it is interrelated to the surrounding conditions, Higher contents of smectite were related to green alder stands and grass heath where moister soils (having in most cases also a finer texture) could be encountered. Green alder need a better-developed soil where they actively contribute to enhanced pedogenesis by stabilising the surface. Grass heath needs a moist environment and contributes to weathering through the release of organic acids.
- Humus forms served as a proxy for the soil biota and soil organic matter composition. The production of chelating organic ligands and acidity was highest at sites having a Eumoder. At these sites, the smectite content was elevated. High percolation rates

and the production of chelating compounds in the soil are believed to promote the development of smectites from mica and chlorite.

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Figure captions

Fig. 1. Location of the Morteratsch proglacial area with isochrones of glacier retreat and position of the topsoil sampling sites (clay mineralogy). Eleven sites were analysed in detail regarding soil chemistry and mineralogy (S1 – 10, A; Mavris et al., 2010). At the other 24 sites, only the mineralogical composition was measured.

Fig. 2. Spatial distribution of the soil and vegetation units in the proglacial area.

Fig. 3. XRD patterns of soil clays (< 2µm) of the soil material close to the glacier front (having an age of 0 yr) and the surface horizon of a soil having an age of 140 yr. The XRD-curves were smoothed and corrected for Lorentz and polarization factors. *d*-spacings are given in nm. EG = ethylene glycol solvation, Mg = Mg-saturation, K = K-saturation and corresponding heating treatments.

Fig. 4. Temporal evolution of the smectite content (mean values and standard error per time zone; no standard error is given if $n \leq 2$) in the clay fraction.

Fig. 5. Relationship between the smectite content and the mica and chlorite content, respectively. For both minerals (mica and chlorite) a trend is observable. The regression equation is given for mica, because only here the correlation is significant.

Fig. 6. Mean concentration (± standard error) of the phyllosilicates in the clay fraction as a function of the vegetation. 0 = no vegetation; 2 = Pioneer grass communities; 3 = Epilobietum fleischeri with single willow shrubs and Alpenrose; 4 = Green alder scrub (Alnetum viridis); 5 = Grass heath on moister soils (Dystric Cambisols, grass species e.g. *Festuca violacea*, *Calamagrostis villosa*, *Phleum rhaeticum*, *Poa alpina*); 6 = Boulder plant communities, partially Epilobietum fleischeri, *Adenostyles leucophylla* and several fern

species; 7 = Rock vegetation (e.g. *Agrostis rupestris*, *Silene rupestris*, *Sempervivum arachnoideum*). No standard error was calculated with $n \leq 2$.

Fig. 7. Mean concentration (± standard error) of the phyllosilicates in the clay fraction as a function of soil organic matter concentration (SOM) and the humus form.

Fig. 8. Mean concentration (± standard error) of the phyllosilicates in the clay fraction as a function of exposure and landform (see also Table 1).

Fig. 9. Phyllosilicate concentrations as a function of the texture and soil thickness.

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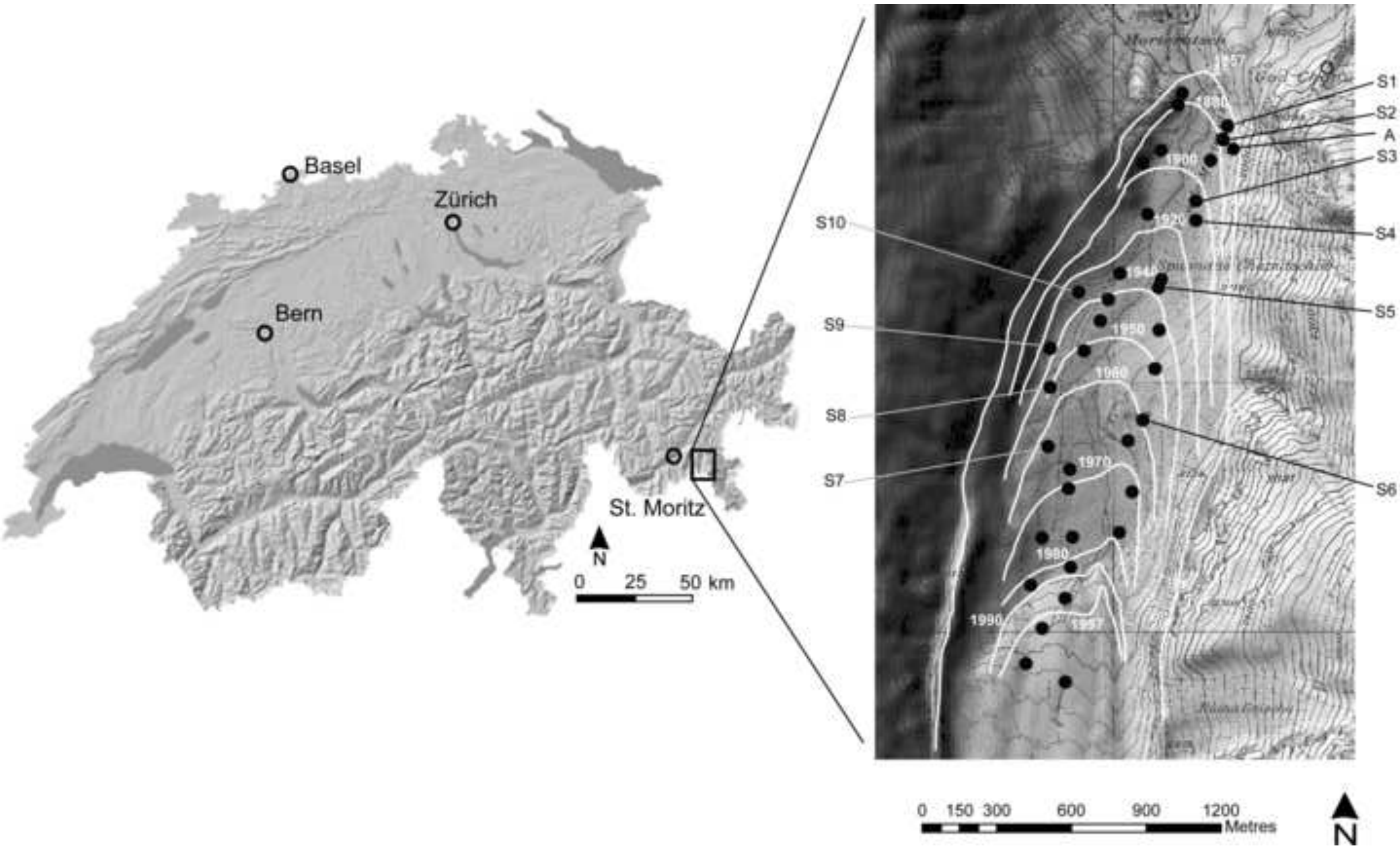

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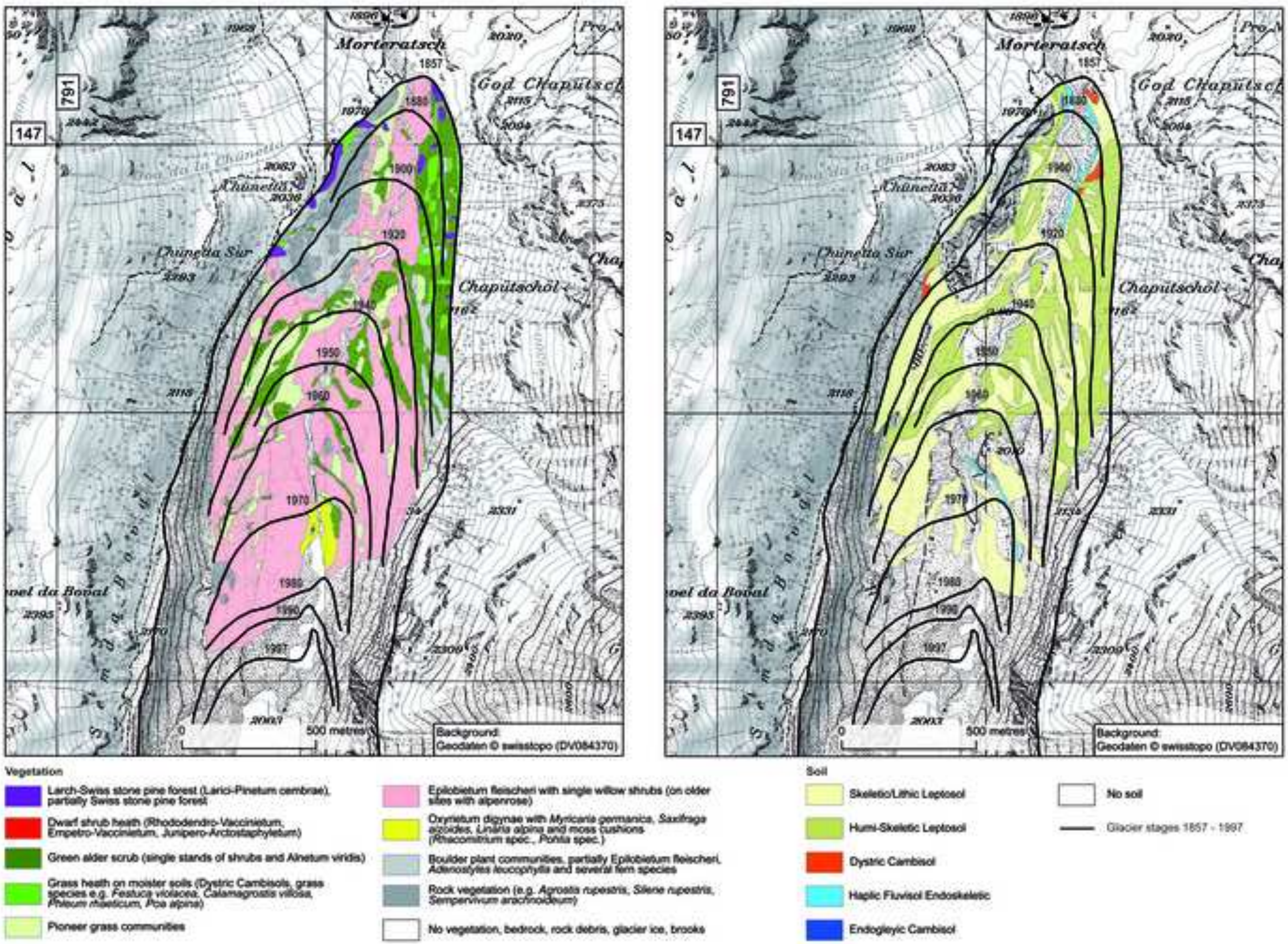

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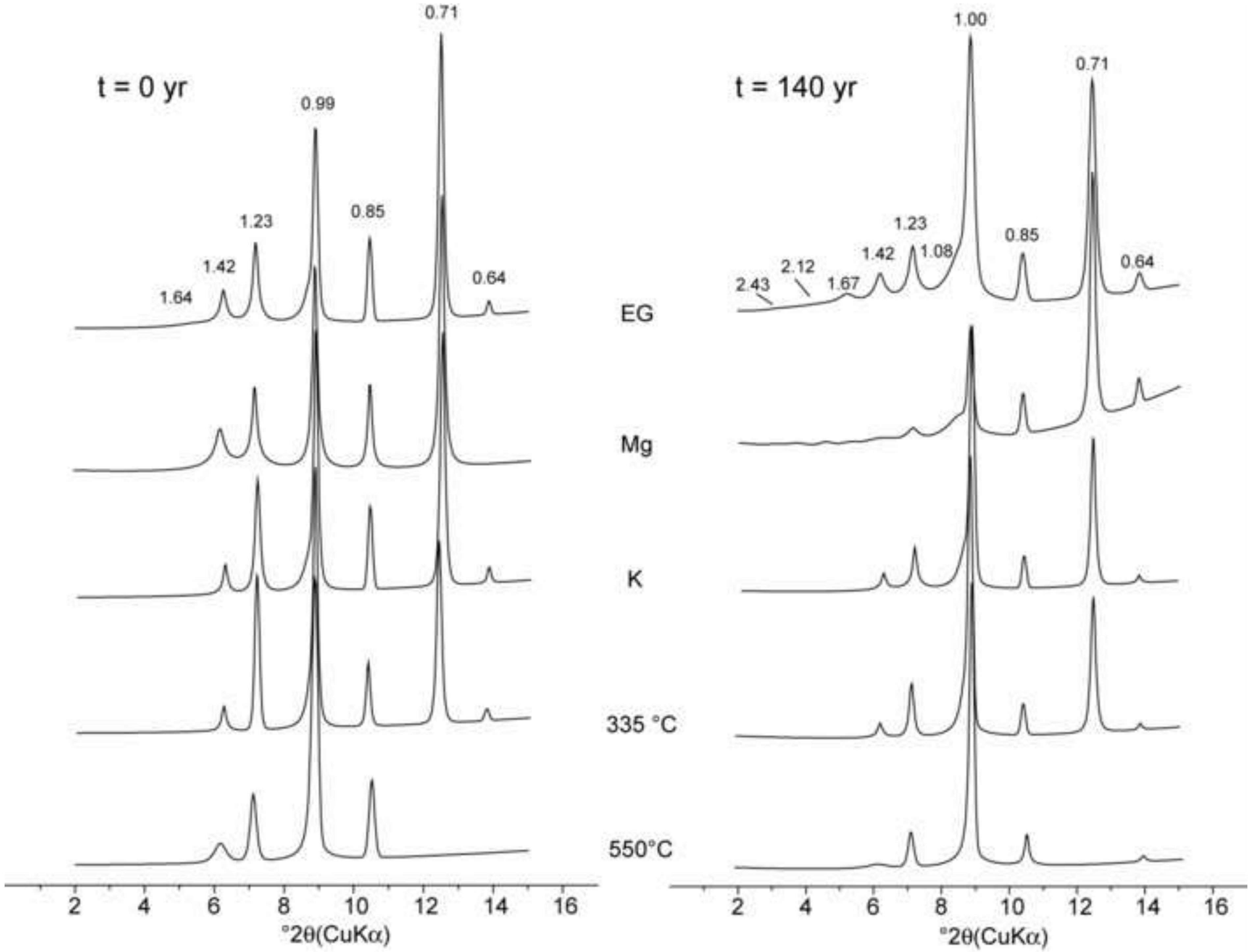

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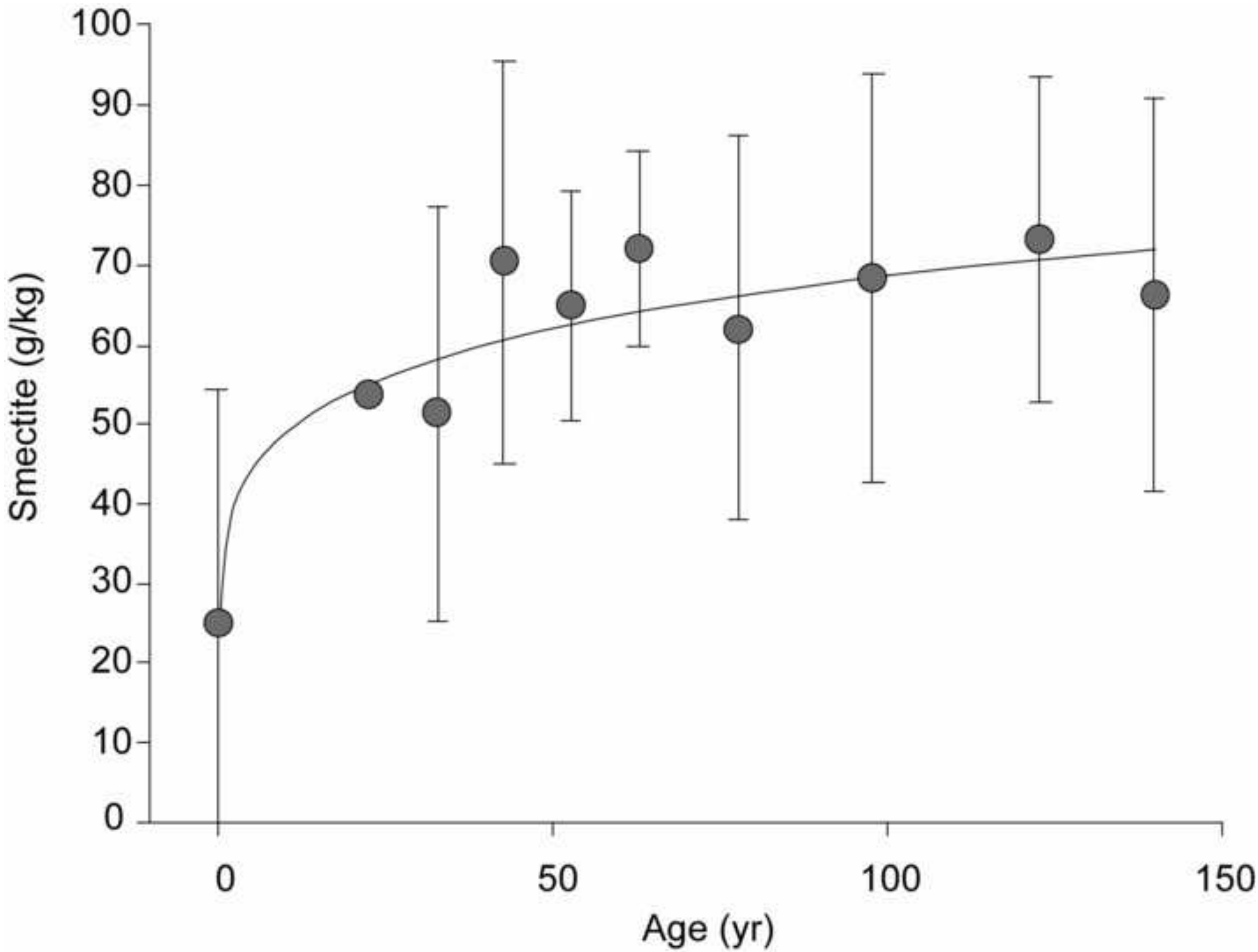

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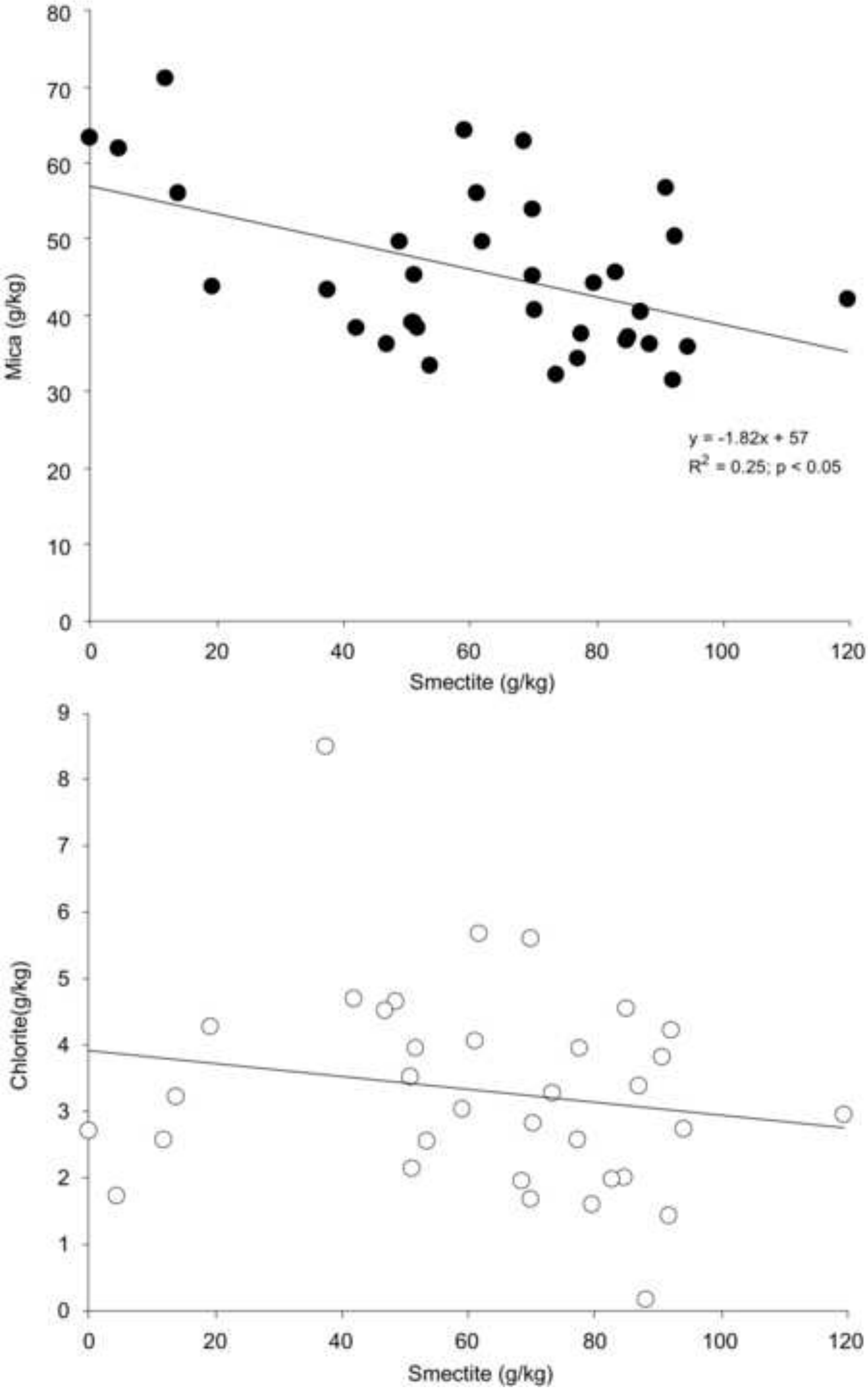

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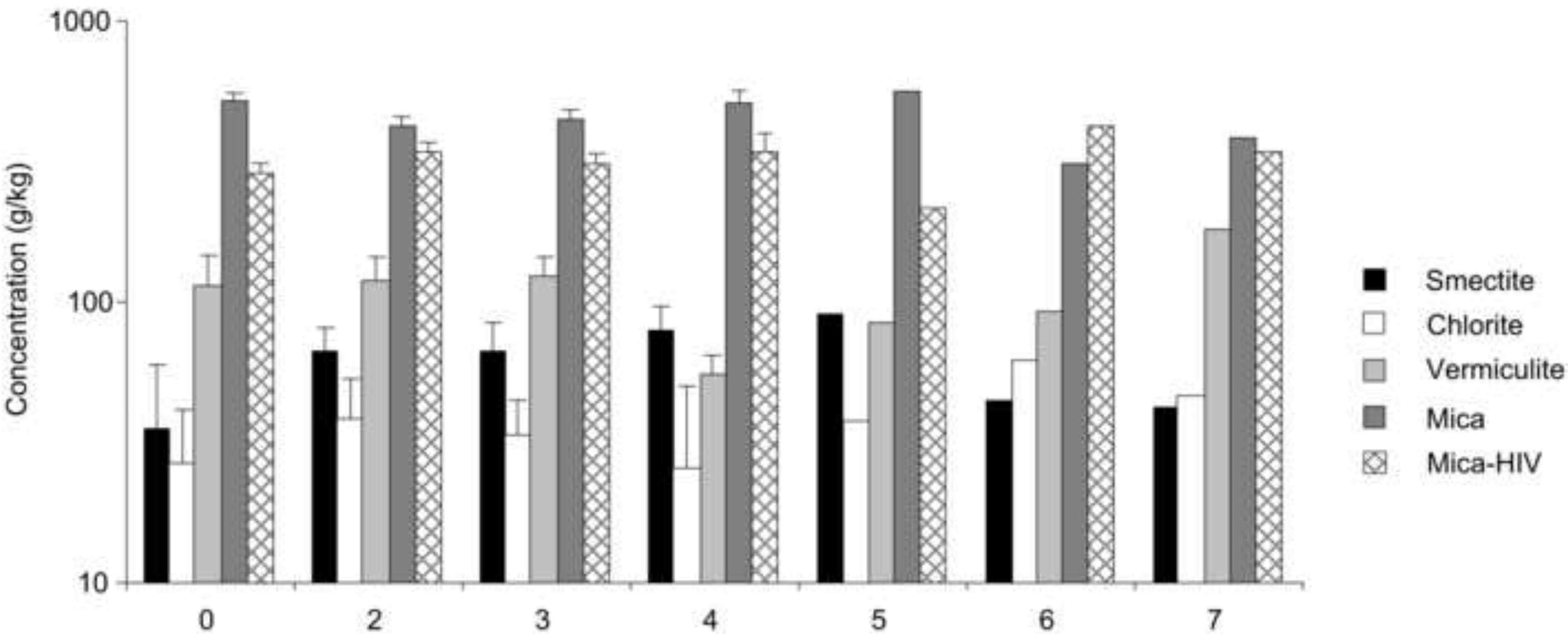

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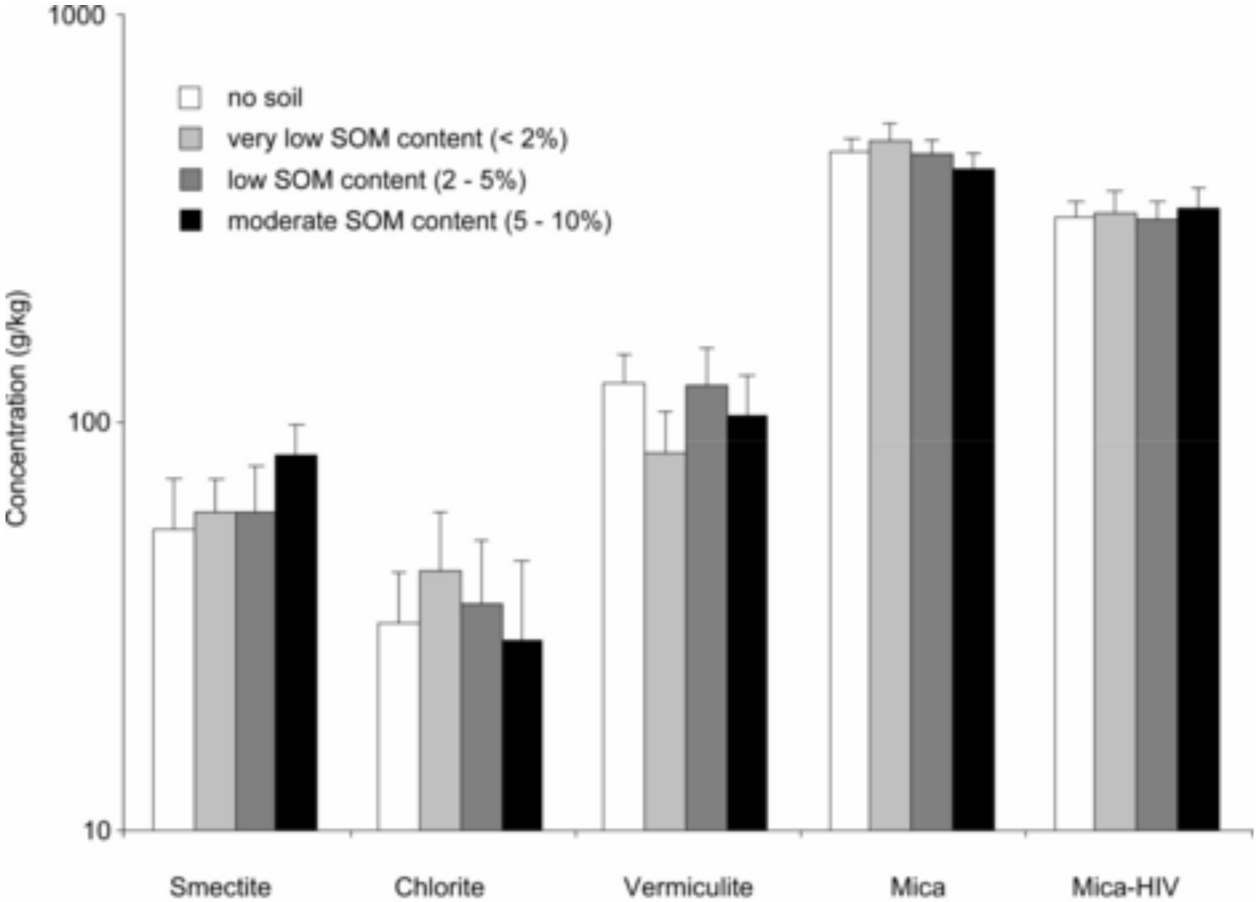

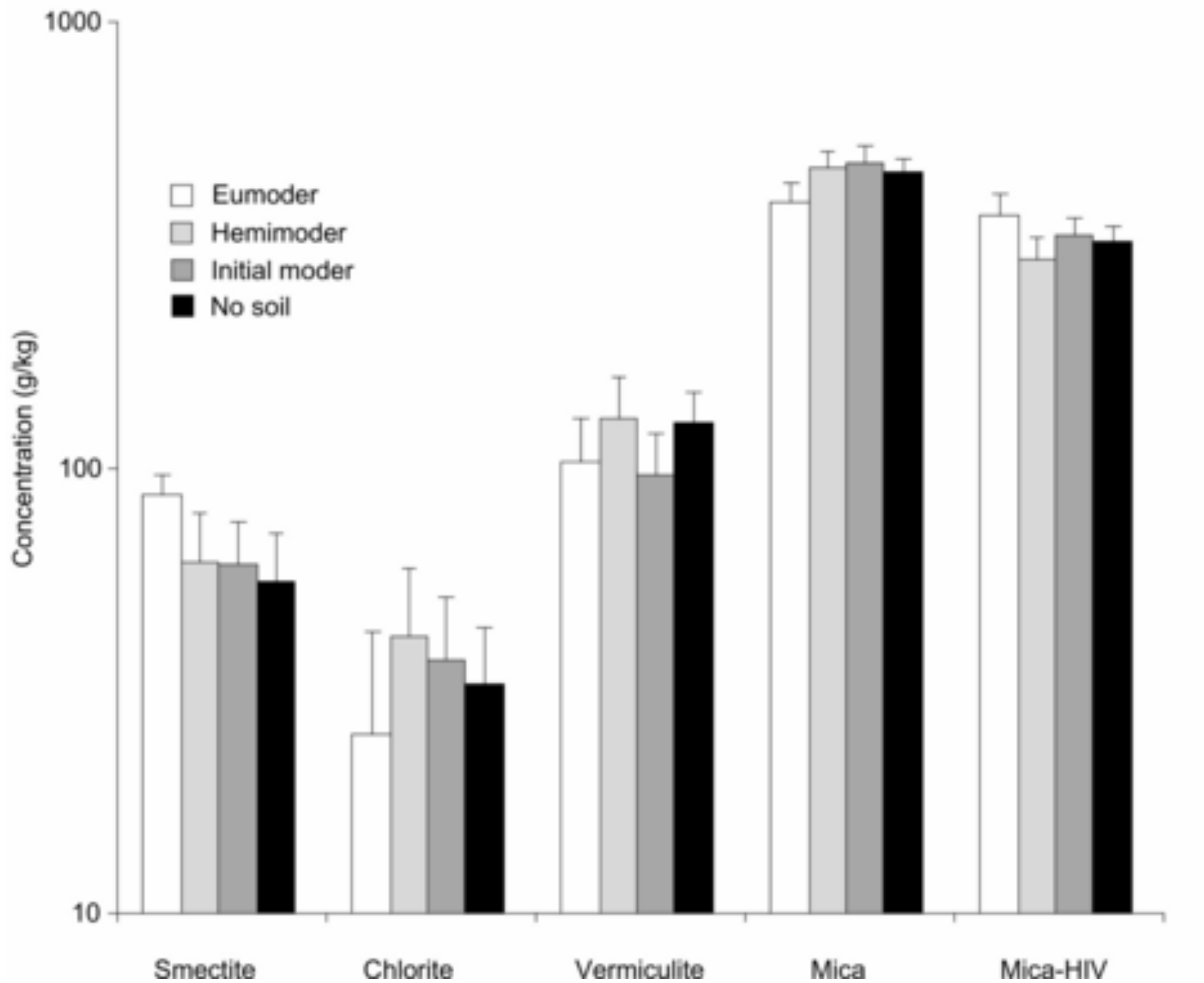

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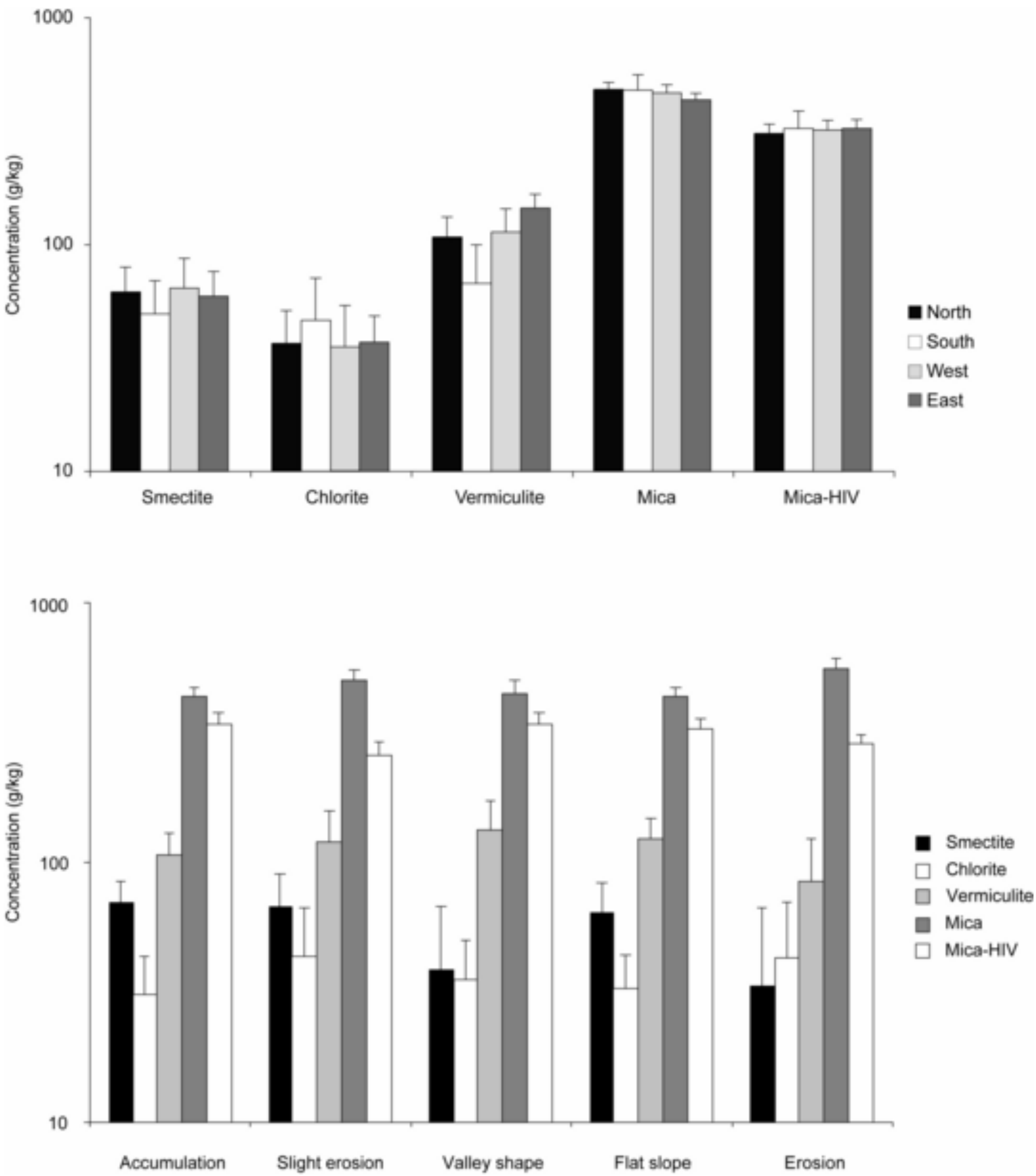

Figure 9
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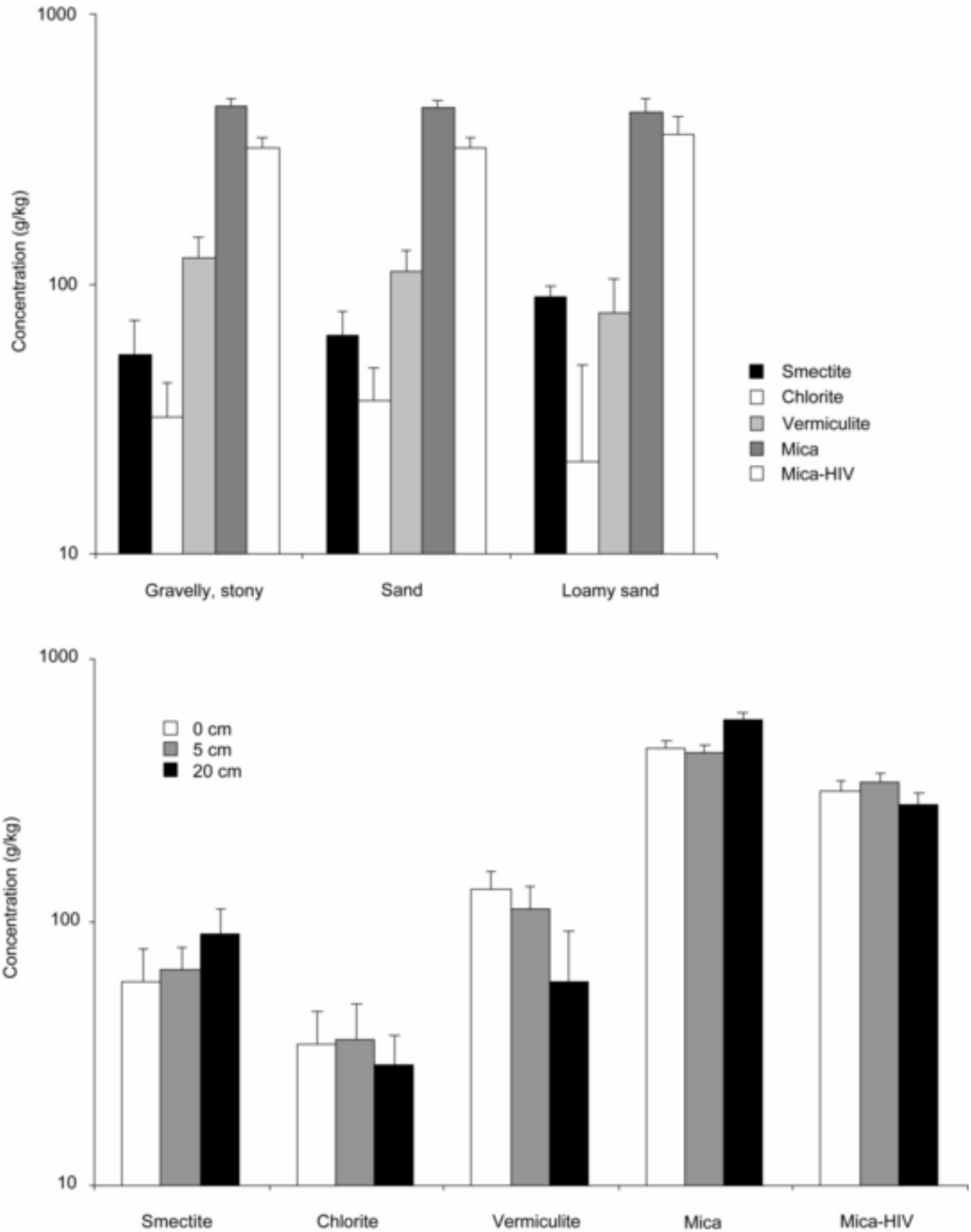

Table 1
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Table 1. Description of landscape forms derived from the digital elevation model

Landscape form	Landscape code	Landscape form: perpendicular to slope (planform curvature)	Landscape form: direction of slope (profile curvature)
Depressions	10	concave	concave
Foot of the slope, flattening slope	20	planar	concave
Flattening slope ridge	30	convex	concave
Valley shape	40	concave	planar
Flat slope	50	planar	planar
Steepening slope ridge	60	convex	planar
Steepening valley	70	concave	convex
Steepening slope	80	planar	convex
Ridges	90	convex	convex

Table 2
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Table 2. Reclassification of landscape forms (in terms of theoretical gains and losses of material)

Landscape form	Code	Landscape code (see Table 1)
Accumulation	12	10 + 20
Slight erosion possible	36	30 + 60
Erosion	89	80 + 90
Flat slope (accumulation and erosion)	50	50
Valley form	47	40 + 70

Table 3
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Table 3. Classification of the slope

Class	Slope (°)
1	0 – 3
2	3 – 6
3	6 – 9
4	9 – 14
5	14 – 19
6	19 – 27
7	27 – 37

Table 4

Table 4. Vegetation units of the proglacial area Morteratsch

ID	Vegetation type	Description
0	No vegetation, bedrock, rock debris, glacier ice, brooks	No vegetation, bedrock, rock debris, glacier ice, brooks
1	Oxyrietum digynae with Myricaria germanica, Saxifraga aizoides, Linaria alpina and moss cushions (Rhacomitrium spec., Pohlia spec.)	*Oxyria digyna, Linaria alpina, Saxifraga bryoides, Cerastium uniflorum, Cardamine resedifolia, Poa alpina, P. nemoralis*
2	Pioneer grass communities.	Geo montani-Nardetum and *Poion alpinae*
3	Epilobietum fleischeri with single willow shrubs and Alpenrose	*Epilobietum fleischeri* with single willow shrubs and Alpenrose
4	Green alder scrub (Alnetum viridis)	*Alnus viridis* and tall perennial herbs, *Salix spec., Poa spec., Deschampsia caespitosa, Avenella flexuosa, Nardus stricta, Festuca spec., Phleum rhaeticum, Anthoxanthum alpinum, Calamagrostis villosa*
5	Grass heath on moister soils	grass species e.g. *Festuca violacea, Calamagrostis villosa, Phleum rhaeticum, Poa alpina*
6	Boulder plant communities, partially Epilobietum fleischeri	*Epilobium fleischeri, Adenostyles leucophylla, Rumex scutatus, Dryopteris spec., Athyrium spec., Gymnocarpium dryopteris, Polystichum lonchitis*
7	Rock vegetation	*Agrostis rupestris, Silene rupestris, Sempervivum arachnoideum*
9	Larch-Swiss stone pine forest (Larici-Pinetum cembrae), partially Swiss stone pine forest	*Pinus cembra, Larix decidua, Rhododendron ferrugineum, Luzula sylvatica, Calamagrostis villosa, Vaccinium spec., Clematis alpina, Homogyne alpina, Oxalis acetosella, Avenella flexuosa*
12	Dwarf shrub heath	Rhododendro-Vaccinietum, Empetro-Vaccinietum, Junipero-Arctostaphyletum